Tectonophysics, 178 (1990) 127–150
Elsevier Science Publishers B.V., Amsterdam – Printed in The Netherlands

Vertical crustal movements along the East Coast, North America, from historic and late Holocene sea level data

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(Received May 23, 1989; revised version accepted November 6, 1989)

Abstract

Gornitz, V. and Seeber, L., 1990. Vertical crustal movements along the East Coast, North America, from historic and late Holocene sea-level data. *Tectonophysics*, 178: 127–150.

Sea-level data from eastern North America are analyzed for evidence of neotectonism, which is inferred from residual sea-level anomalies, after removal of glacial isostatic and eustatic components. The sea level is examined on two different time scales, using long-term Holocene paleosealevel indicators, and short term (≤ 100 yrs) tide-gauge records.

The major process affecting Holocene sea levels in this region is the continuing response to glacial unloading. The northward migration of the collapsed bulge peripheral to the retreating ice sheets is clearly delineated by the displacement of the hingeline between glacioisostatic subsidence and uplift from around 43° to 48°N, over the last 14,000 years. Similarly, the zone of maximal subsidence (collapsed forebulge) has moved from ~ 36°–42°N prior to 11,000 years to 44°–46°N during the last 5000 years.

Tide-gauge data for the East Coast covering the last 100 years indicate a sea-level rise of around 2–4 mm/yr. Subtraction of the late Holocene trend reduces the regional mean from 2.72 ± 0.71 mm/yr to 1.26 ± 0.78 mm/yr, which is close to the value (1.47 ± 0.68 mm/yr) obtained by using viscoelastic model corrections (after Peltier, 1986). Thus, nearly half of the recent regional sea level rise appears to be of glacioisostatic origin. The corrected mean regional trend lies within a few tenths of a mm/yr of the estimated mean global eustatic trend (~ 1.1 mm/yr; Gornitz and Lebedeff, 1987).

Generally, the results show no evidence for major tectonic movements, especially for large fault offsets. Nevertheless, several anomalous areas persist, that are suggestive of gentle warping, after glacioisostatic and eustatic processes have been taken into account. These correspond to areas with known Neogene epeirogenic movements, and include the region between Savannah, Georgia–Charleston, South Carolina (subsidence), Cape Fear Arch (uplift), southern Chesapeake Bay (Late Pleistocene–Holocene uplift, succeeded by recent subsidence), Montauk and the Long Island platform (uplift?). Subsidence in the Charleston area and southern Chesapeake Bay may be in part caused by groundwater withdrawal.

Introduction

Relative vertical movements of coastal regions can occur over a broad range of wavelengths. Flexure of the lithosphere, due to differential loading, can produce broad warping of the passive continental margin, with wavelengths of many hundreds or thousands of kilometers. Examples include the removal of the Fennoscandian and Laurentide ice sheets from the continental lithosphere (Peltier, 1987; 1986), or loading of the continental margin by sediments and water. Post-rift cooling and densification of oceanic lithosphere also contribute to subsidence of the continental margin (Heller et al., 1982). Changes in relative land elevation can also occur over shorter

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distances (< 100 km); these are likely to stem from tectonic deformation and fault displacements within the crust.

The purpose of this paper is to analyze sea level data from eastern North America for indications of neotectonic deformation. Our approach is to consider as signal the residual sea-level anomaly after glacioisostatic and eustatic components are accounted for.

Two kinds of sea level data are utilized here:

(1) Long-term paleosealevel indicators for the Holocene are derived from ^{14}C-dated peat, shells, corals, wood, etc. (Pardi and Newman, 1987).

(2) Short-term sea-level data ($\leq$ 100 yrs) are obtained from tide gauges (Lyles et al., 1987; Permanent Service for Mean Sea Level, Bidston Observatory, England).

Long-wavelength changes in elevation associated with ice loading can be modeled by assuming a viscoelastic earth (Peltier, 1987, 1986). The model accounts for gravitational interactions among ice sheets, land and ocean. Since the completion of glacial melting about 6000 yrs B.P., the rate of eustatic sea level rise has decreased sharply. However, the model predicts that the glacioisostatic recovery is not yet complete, in many places. Thus, late Holocene ($\leq$ 6000 yrs) sea level curves largely reflect the continuing glacioisostatic response (Clark et al., 1978; Peltier, 1987). Subtraction of the late Holocene sea level trends from the tide-gauge trends should minimize long-wavelength vertical motions including the glacial isostatic effects (Gornitz et al. 1982; Gornitz and Lebedeff, 1987). However, short-wavelength movements may still persist, even after such filtering.

In this study, we examine the regional variability in sea level records over the last 100 years from tide-gauge records and over the Holocene from stratigraphic data. The sea level data are processed to emphasize short-wavelength differences. These differences will be related to geologic structures and seismicity along the Atlantic coast of North America. The present methods, however, are inadequate to resolve long-wavelength tectonic deformation.

Over the last 100 years, rates of relative sea level rise along the East Coast, including the Canadian Maritimes, range between 2 and 4 mm/yr (Gornitz and Lebedeff, 1987). These values lie well above the estimated global eustatic sea level rise of ~ 1–1.5 mm/yr (Gornitz et al. 1982; Barnett, 1983, 1984; Gornitz and Lebedeff, 1987). Several processes can produce anomalous rates of sea level rise. These include neotectonism, sediment loading, subsidence of the continental margin, or a migrating forebulge collapse (Walcott, 1972). The relative contributions of these various processes will be assessed below.

Several papers have also interpreted the recent sea level records of eastern North America in terms of neotectonic processes. Braatz and Aubrey (1987) find that the observed spatial variation in sea level trends corresponds to still-active Mesozoic–Cenozoic structural features. On the other hand, Uchupi and Aubrey (1988) infer a weak link between a relative sea level change and Paleozoic suspect terranes.

In general, the seismicity in intraplate areas is much more subdued than along plate boundaries. Nevertheless, the Atlantic seaboard of North America is characterized by a relatively high level of seismicity, concentrated in well defined clusters (Fig. 1). In some cases, these are associated with known structures, such as the Newark Basin in southern New York and northern New Jersey (see fig. 2 in Seeber and Armbruster, 1988). The earthquakes are believed to occur along pre-existing faults, reactivated in the neotectonic regime (Ratcliffe, 1980; Seeber and Dawers, 1989).

In spite of rapid progress during the last decade, several fundamental problems concerning seismicity, deformation, and structure along the Atlantic Seaboard remain unresolved. While seismicity levels suggest significant rates of neotectonic deformation, this has not been substantiated by geologic evidence. In the few instances where Neogene sediments are faulted, offsets are small (e.g., Prowell, 1988). Moreover, none of the large historic earthquakes have been associated with surface ruptures. A relationship of specific earthquakes with mapped faults has generally been lacking. Recently, accurate hypocenters and detailed mapping establish such a connection (e.g., the October 19, 1985 earthquake in Westchester Co., New York and the Dobbs

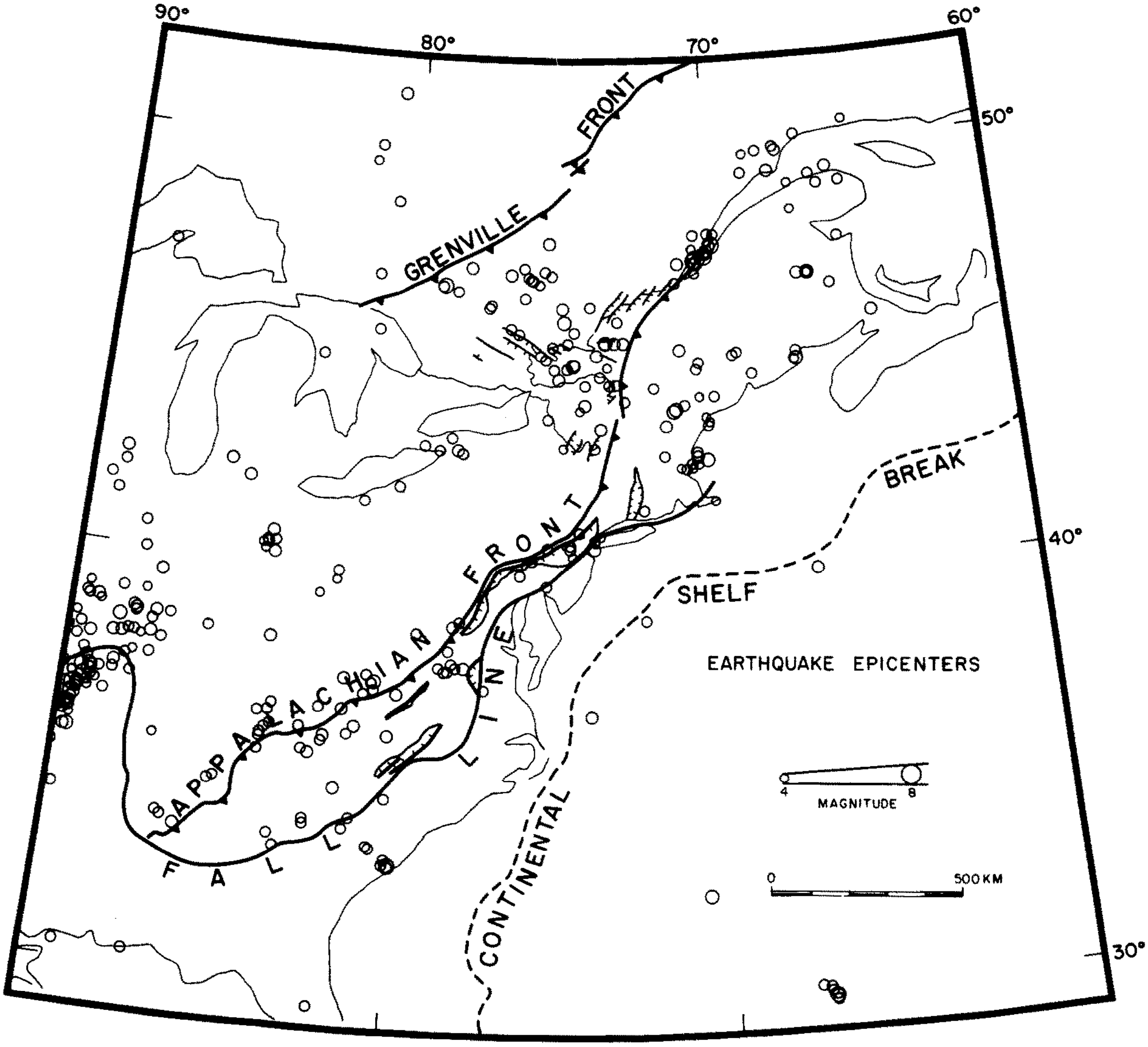


Fig. 1. Earthquake epicenters in eastern North America, showing magnitude greater than or equal to 4.0. Data are taken from the Electric Power Research Institute, Catalog of Central and Eastern North American Earthquakes to 1985 (J. Carl Stepp, pers. commun., 1989). Major geologic structures are from P.B. King, 1969, Tectonic Map of North America, 1 : 5,000,000.

Ferry fault, 10 km north of New York City; Seeber and Dawers, 1989). The seismogenic faults are reactivated minor structures with a multistage evolution, but with very little total displacement. Seismogenic faults with substantial Neogene displacement, however, cannot be ruled out from the data available so far. Holocene sea-level data could provide independent information on faults with large slip rates, if they exist.

The irregular correlation between seismicity and structure may be the result of temporal clustering with time constants much longer than the historic period (Seeber and Armbruster, 1988). According to this hypothesis, a systematic correlation between seismicity and structure should be detected, if sampled over a long enough period. Different parts of the seismic belt could remain active for relatively short periods, and could therefore be missed during the short historic period. Seismicity could suddenly start, possibly with a large earthquake, in an area with no previous record of historic seismicity. The relevance of this hypothesis to earthquake hazard makes the testing of this hypothesis an urgent issue. A comparison between

Holocene sea level data and historic tide data could contribute to such a test.

Procedures

Data bases available include tide-gauge measurements (Lyles et al. 1987: data through 1986) and Permanent Service for Mean Sea Level, Bidston Observatory, U.K. (through 1983) and a compilation of ^{14}C-dated paleosealevel indicators (Pardi and Newman, 1987). The space–time distribution and limitations of the ^{14}C data base are described in Pardi and Newman (1987). Sources of error include uncertainties in estimating past tidal ranges, the relation of the dated material to sea level position, and dating errors. Errors in ^{14}C dating can arise from incorporation of sea water with an "old" apparent age into marine organisms, recrystallization of coral and aragonitic molluscs, in a non-closed system; and contamination by living rootlets, and humic acids from modern plants or old, dissolved carbonates, in groundwater (Sutherland, 1987). Many of these sources of error can be avoided by careful field sampling and pre-treatment of specimens prior to analysis, and by checking against other methods of dating, e.g., pollen analysis. Systematic differences related to the nature of the material dated were not found in the present study. Shells are often less reliable sea level indicators than basal salt marsh peats, because of their mobility and ocurrence over a fairly wide depth range. Yet, in many cases, shells were found to fall on the same curve as other sea-level indicators, while, conversely, some peat

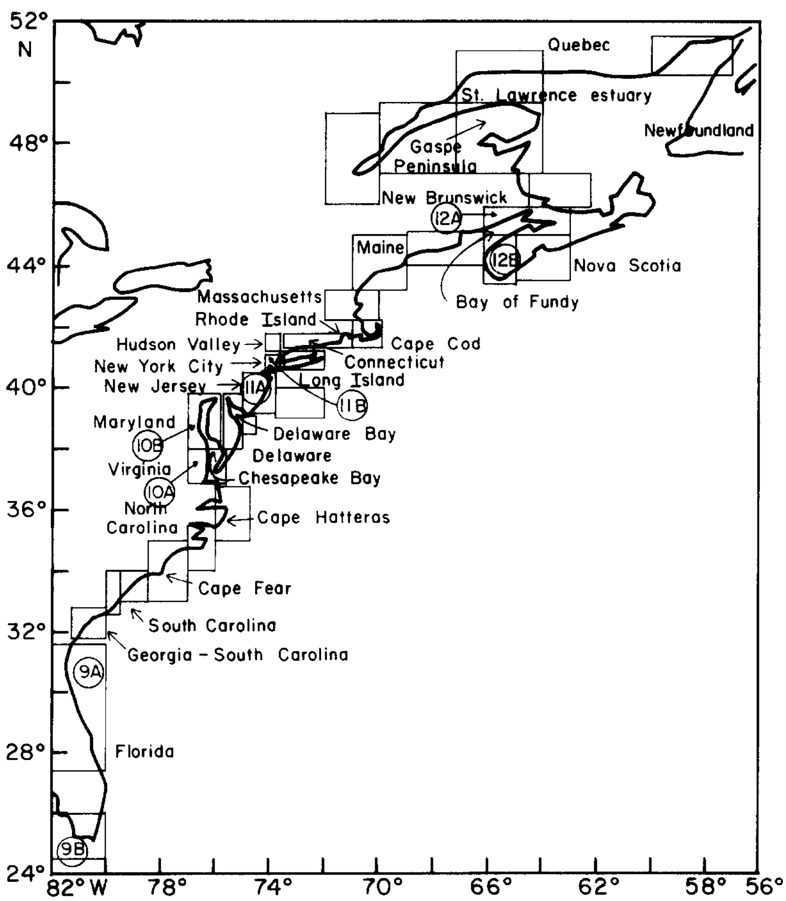


Fig. 2. Index map of eastern North America showing location of cells from which ^{14}C-derived paleosealevel trends have been obtained. Circled numbers refer to figures.

samples were occasionally way off. Methods of processing the sea level data are further discussed in Gornitz and Lebedeff (1987).

Contour plots of Holocene data, grouped into 2500-yr intervals provide a synoptic view of the changes in shore position over time from southernmost Florida to Nova Scotia. Paleosealevel positions for points between Key West, Florida and Newfoundland, Canada (24°–42° N, 65°–85° W and 42°–60° N, 55°–72° W), were projected normal to the digitized coastline, after smoothing out bays and peninsulas, starting at Key West, in a series of 1000 year increments, between 0 and 14,000 years B.P. Points were also projected normal to latitude. Because few differences occur between both sets of plots, the results are described in terms of the latitudinal projections. These curves present a detailed summary of the deformation along the shoreline in both space and time.

The Holocene ^{14}C paleosealevel data have also been grouped into cells, small enough to have undergone a fairly uniform change in level, yet large enough to enclose a sufficient number of data points to generate reliable sea level curves. These cells are more or less contiguous along the eastern seaboard and offshore (Fig. 2). A long-range trend has been calculated from the paleosealevel data for each cell. In general, only the last 6000–7000 years of data have been used, in order to avoid the period of rapid sea level rise following deglaciation. Depending on the data, least squares linear regression or higher order polynomial curves have been fitted to the data. The results from the Holocene sea level data will be compared with results from tide-gauge records.

The rates of Holocene sea level change derived from the ^{14}C data include both glacioisostatic and other flexural components. The viscoelastic earth model (Peltier, 1986, 1987) isolates the purely isostatic component. The relative sea level history predicted by the geophysical model is compared to the observed ^{14}C data, in order to constrain the lower-mantle viscosity. The model largely accounts for the observed relative sea level data. Therefore, the results obtained from both methods are not completely independent, and can be expected to agree, in most cases. However, the relative sea level curve at any given site is not exclusively the product of glacial isostatic adjustment; epeirogenic, tectonic and other loading effects may also be involved. Thus, where large differences in long-term trends are observed, and if errors in the ^{14}C data base can be discounted, the presence of non-glacioisostatic vertical movements may be responsible.

The tide-gauge data have been corrected by subtracting both the estimated recent eustatic sea level rise (~1.1 mm/yr; Gornitz and Lebedeff, 1987), and the long-term component obtained from both the ^{14}C data, calibrated for past atmospheric ^{14}C fluctuations (Gornitz and Lebedeff, 1987), and from the viscoelastic model (Peltier, 1986). The residuals are interpreted in terms of neotectonic movements, and other potential causes of recent crustal deformation.

Results

Survey of Holocene shoreline changes

Latitudinal projections of paleoshorelines

Figure 3 represents latitudinal projections of paleoshorelines between 24° and 60° N, in 1000-year increments, to 14,000 yrs B.P. Prior to 11,000 yrs B.P., a sharp boundary, or hingeline, was located near 43°–44° N (Fig. 3a–d). This boundary separated points to the south, lying below present sea level, from points to the north, above sea level. Between 8000–11,000 yrs B.P., the hingeline (~45° N) and a zone of enhanced subsidence to its south (36°–42° N), migrated north (Fig. 3d–f). During this period, paleosealevels south of the hingeline were as deep as −70 to −80 m. Data are scarce between 6000–7000 yrs B.P. (Fig. 3g,h). By 6000 years B.P. or less, the hingeline had shifted north to ~48° N (Gaspé–Quebec, Canada, Fig. 2h–m). Maximum subsidence, south of the hingeline, occurs at around 45° N.

Paleosealevel contours

Because of discontinuities in spatial and temporal data coverage, data points, were only contoured for selected areas, described below, at 2500-year intervals.

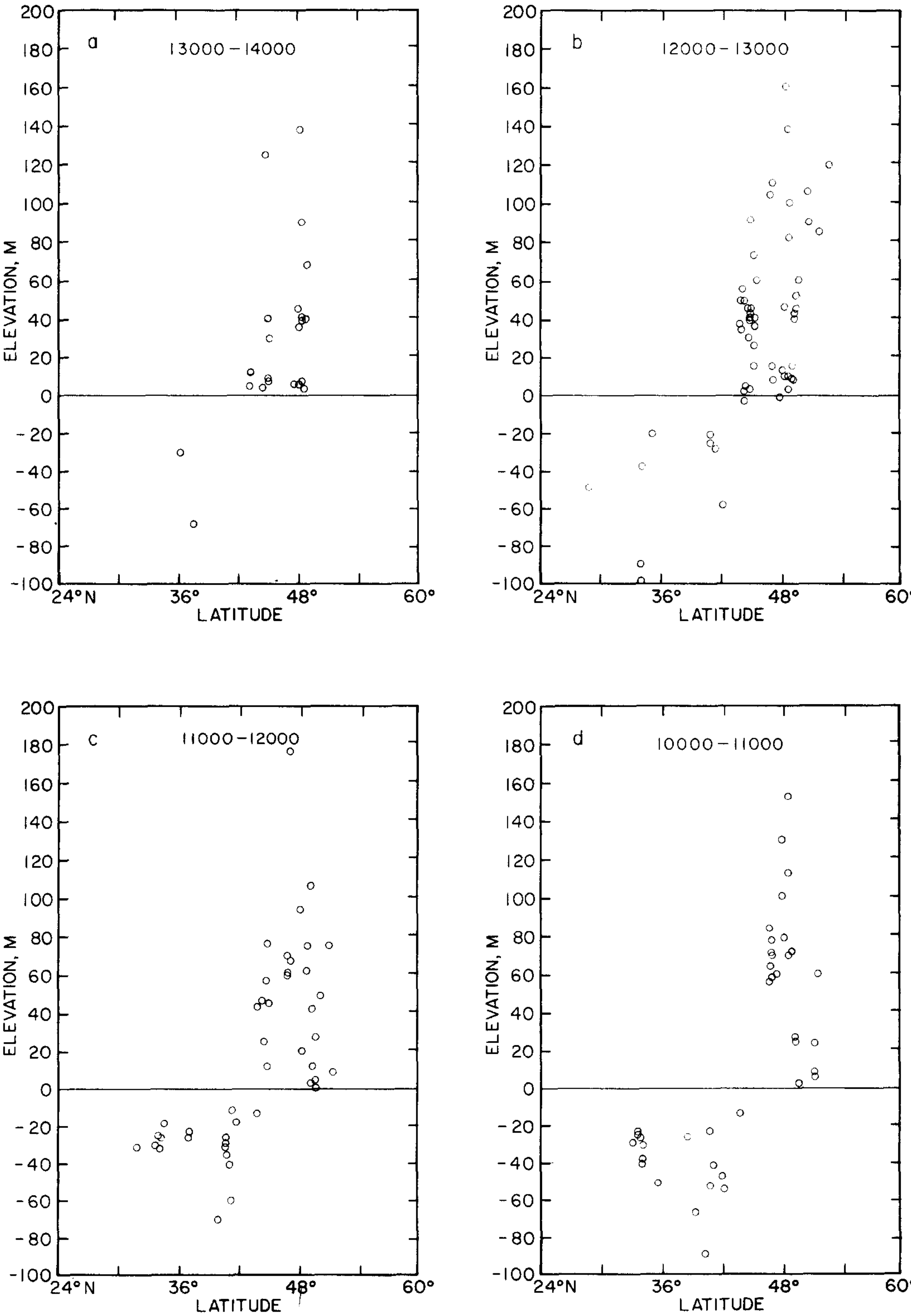


Fig. 3. Latitudinal projections of paleoshorelines from Key West, FL (24° N) to Labrador (60° N), in 1000-year intervals.

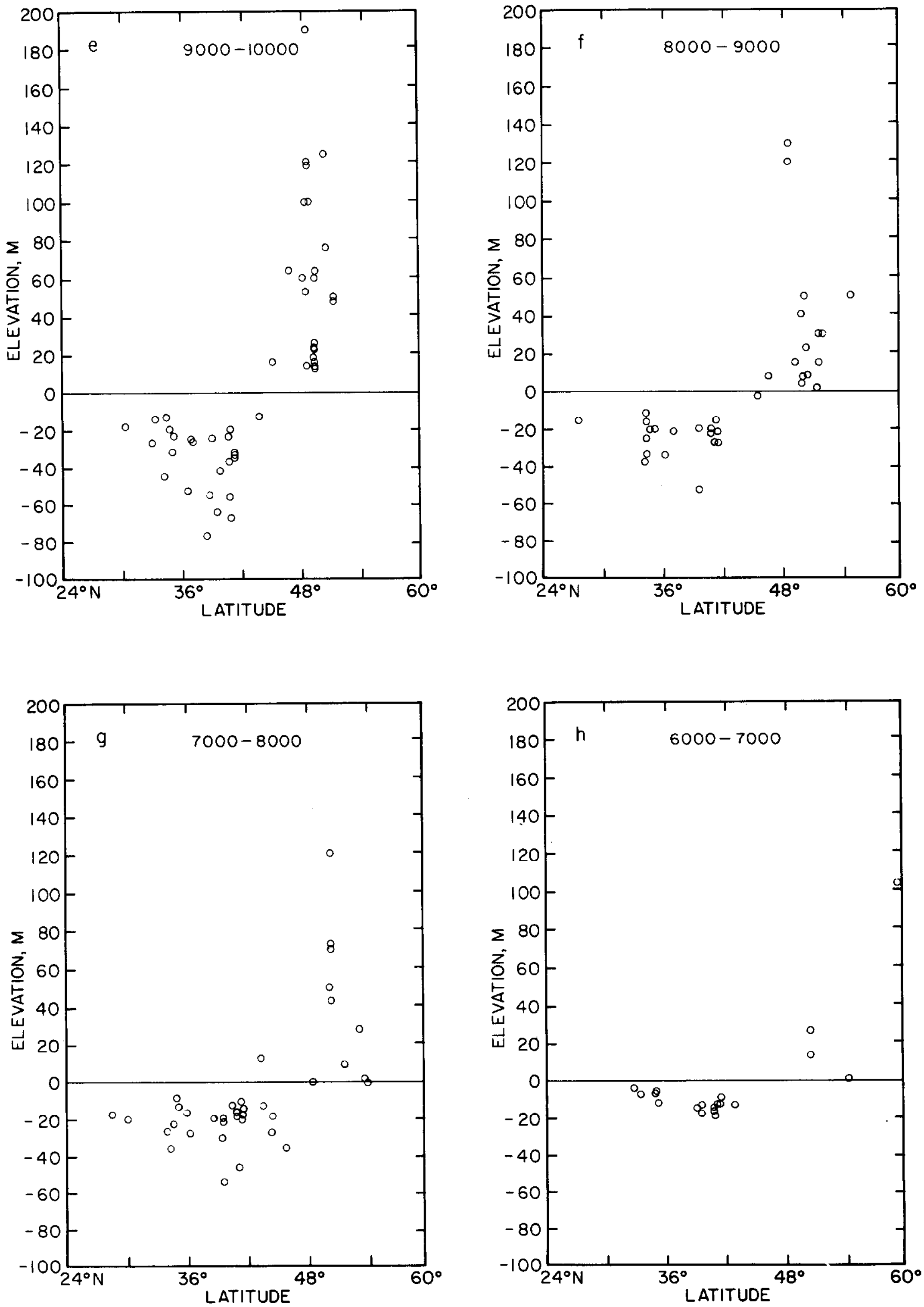
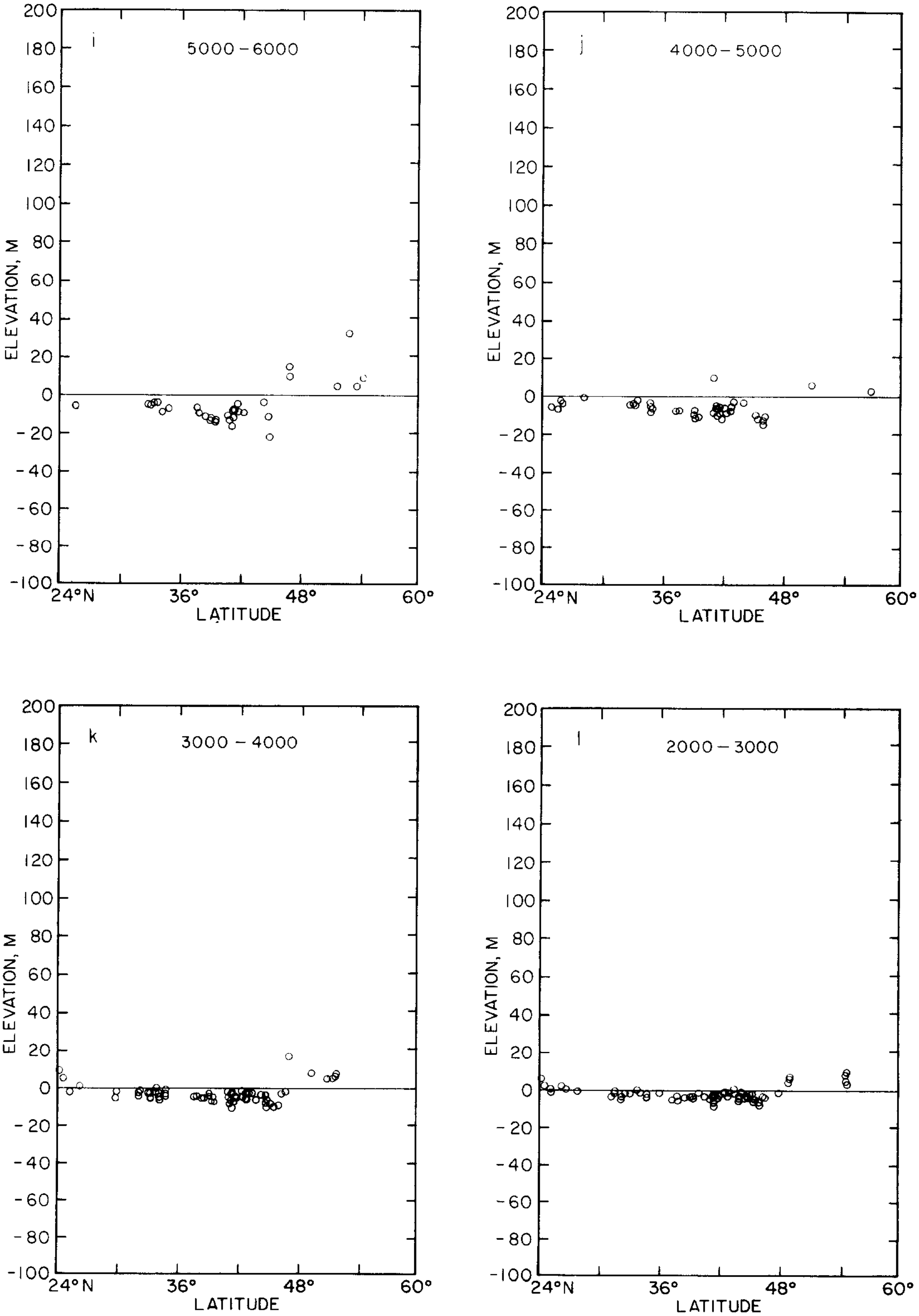


Fig. 3 (continued).



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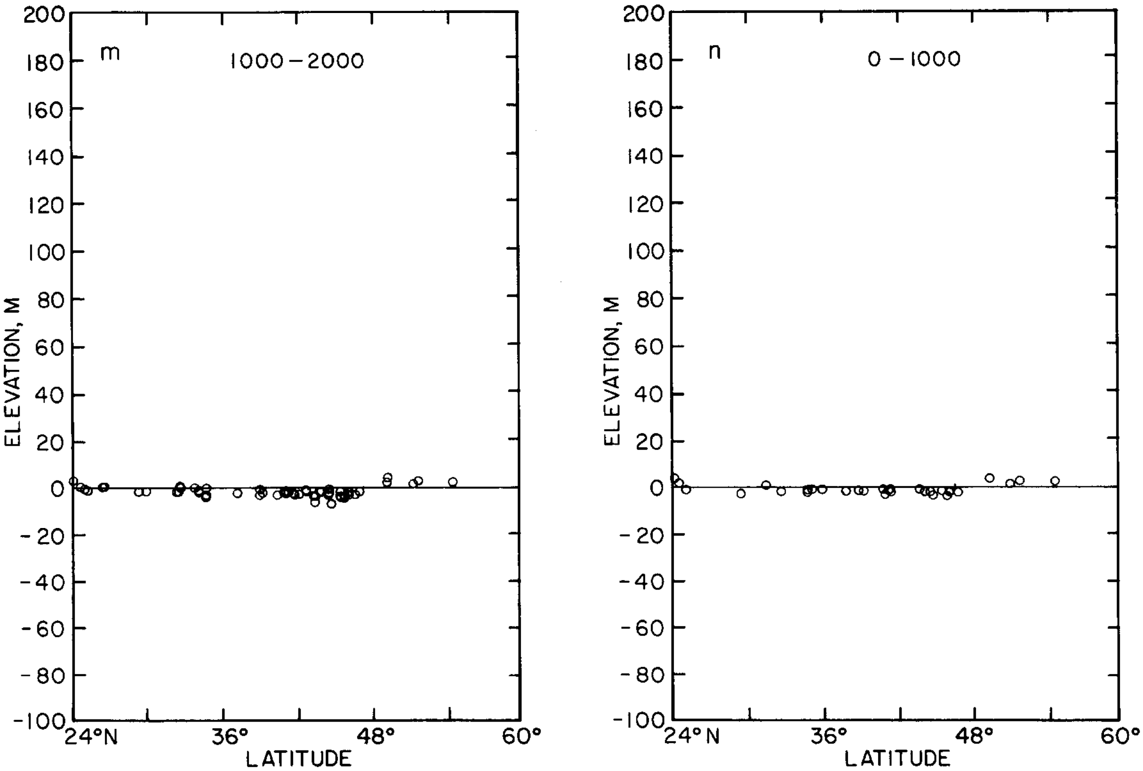

Fig. 3 (continued).

Cape Hatteras–Chesapeake–Delaware Bays. Over the course of the last 7500 years, the upper or inland parts of Chesapeake Bay and Delaware Bay have subsided somewhat more than the adjacent coast or offshore area (Fig. 4). Furthermore, the regional paleosealevel tilt is north-northeast toward Delaware Bay and New Jersey. The differential subsidence may be attributed to sedimentation or to gentle late Pleistocene–Holocene crustal movement (see below).

New York–New Jersey–Connecticut. Within the last 5000 years, the middle Hudson Valley has subsided several meters deeper than elsewhere on the Long Island Platform (Fig. 5). However, prior to 5000 years B.P., the locus of maximum subsidence was farther south, in the New York City area. Possible explanations include the northward migration of the collapsed forebulge, and/or renewed faulting in the Hudson Valley, north of New York city (Newman et al., 1987).

Gulf of Maine Platform–Bay of Fundy–Nova Scotia. Between 0–7500 years B.P., subsidence has concentrated along the upper Bay of Fundy, and around Amherst, Nova Scotia (Fig. 6). Two dates older than 7500 years B.P., are above present sea level. This region shows initial isostatic rebound or emergence, followed by later submergence. This is a transitional zone, located at the edge of the former ice sheet, lying between a zone of continual emergence, where land was previously glaciated (Region I), and a zone of continual submergence, due to the collapsing forebulge (Region II; Clark et al., 1978). In the Bay of Fundy–Nova Scotia region, this corresponds to zones B and C (Scott et al., 1987).

Recent and late Holocene sea level trends

Tide-gauge data

Tide-gauge data along the East Coast are summarized (Table 1, col. 1). The regional mean sea

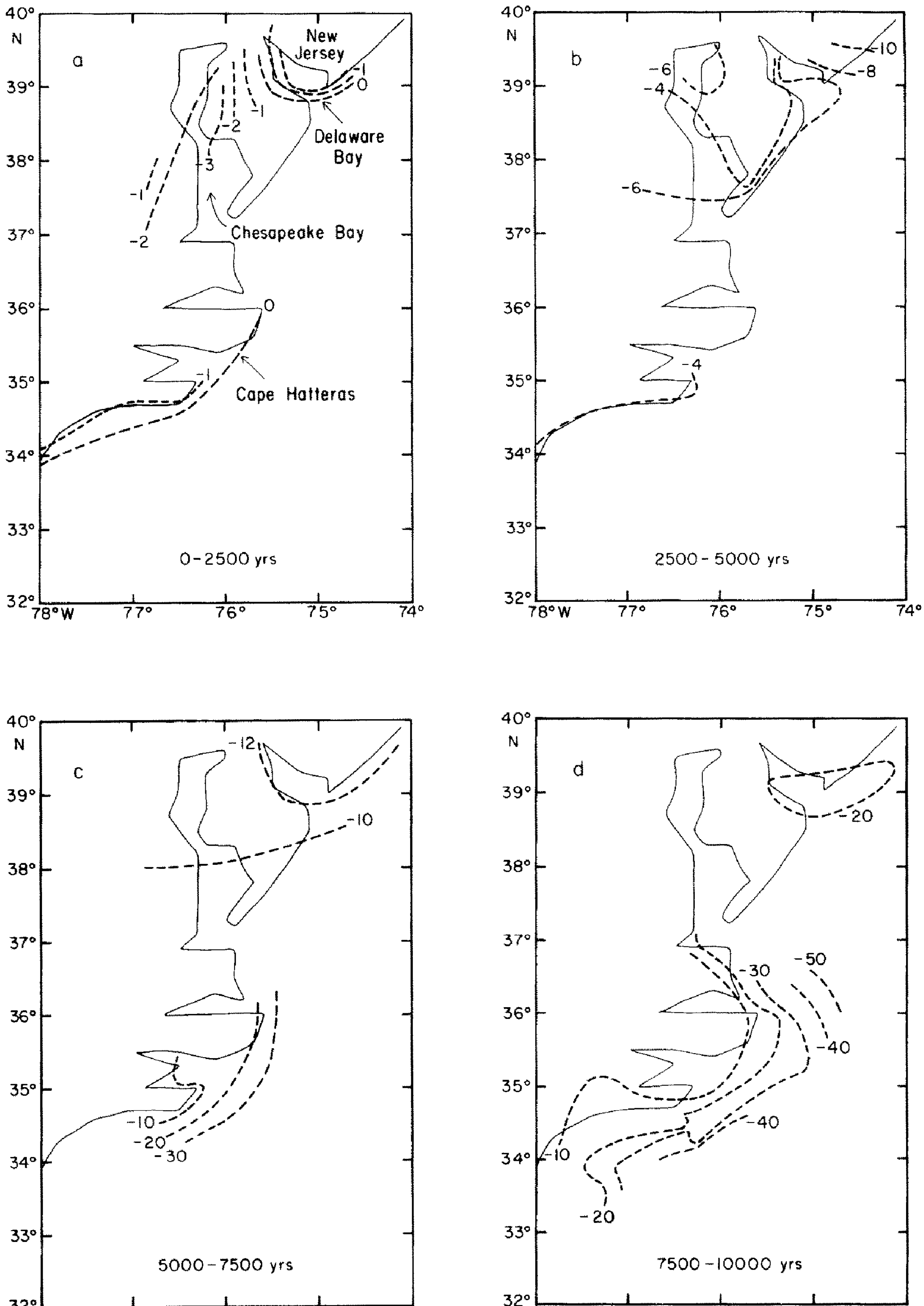

Fig. 4. Paleosealevel contours, 2500-year intervals, Cape Hatteras, Chesapeake Bay and Delaware Bay. Contour intervals for Figs. 4–6 are in meters.

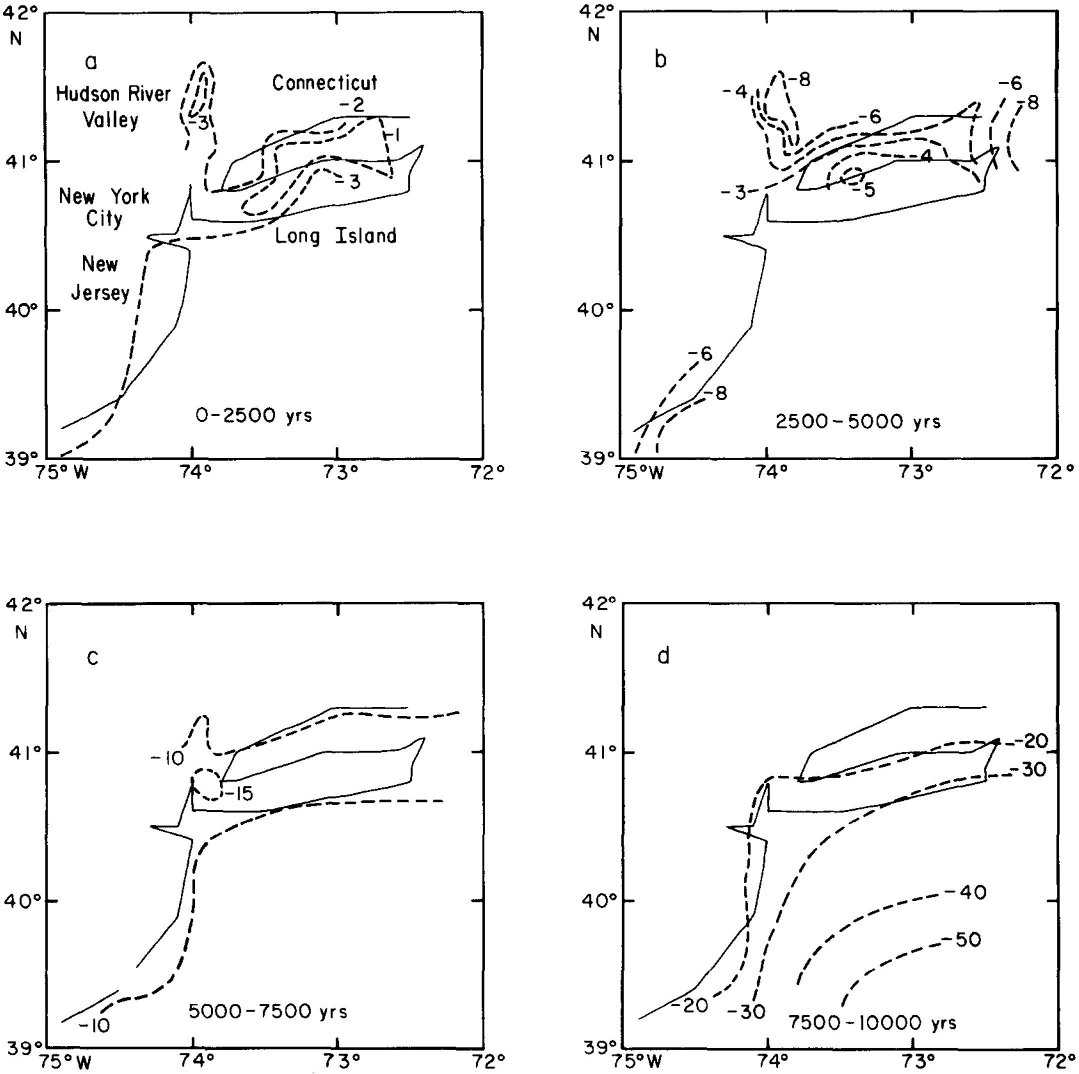

Fig. 5. Paleosealevel contours, 2500-year intervals, New Jersey, New York and Connecticut.

level rise for 40 stations between Key West, Florida and St. John's, Newfoundland (excluding Harrington Harbor and Buzzards Bay) is 2.72 ± 0.71 mm/yr (standard deviation). However, the spatial variation is non-uniform, with higher than average trends concentrated between Chesapeake Bay and New Jersey, and around Halifax, Nova Scotia (Fig. 7). *

Along the southeast Atlantic coast, sea level (SL) trends for stations between Key West, Florida and Wilmington, North Carolina range between 1.8–3.4 mm/yr. The SL trend for Wilmington, North Carolina (1.8 mm/yr) is somewhat less than those of adjacent stations, which may reflect continuing uplift along the Cape Fear arch (Markevich, 1985; Braatz and Aubrey, 1987).

Stations between Portsmouth, Virginia and Portland, Maine show a northward decrease in the rate of SL rise. Recent SL trends for southern Chesapeake Bay are anomalously high (see below). Above-average trends along the New Jersey coast could be due, in part, to sediment compaction, or to groundwater withdrawal (Davis, 1987). Buzzards Bay, Mass. is anomalously low (0.52 mm/yr),

* See Fig. 8 for location of tide-gauge stations discussed in text.

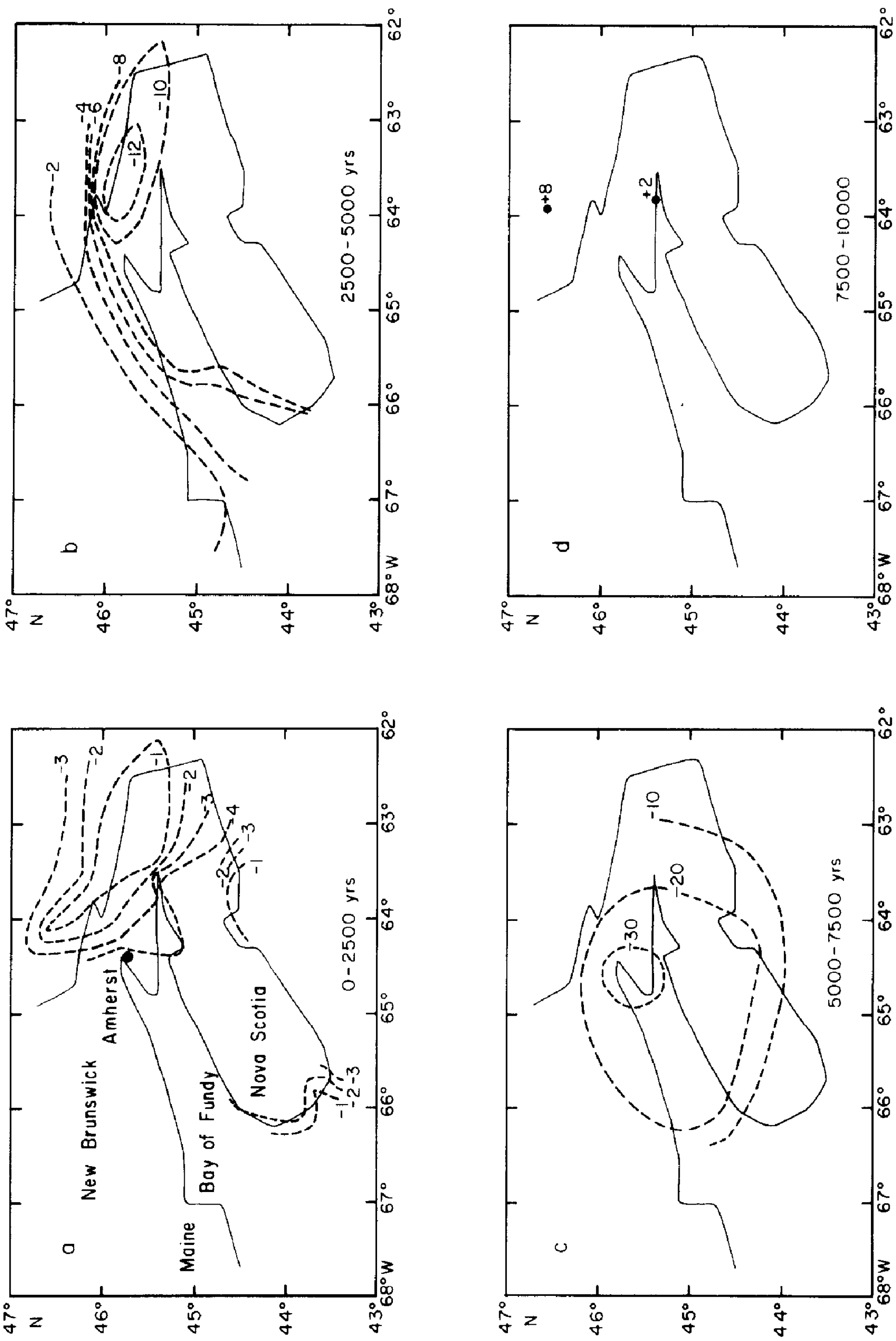

Fig. 6. Paleosealevel contours, 2500-year intervals, Maine, Nova Scotia and Bay of Fundy.

TABLE 1
Sea level data of eastern North America (mm/yr)

	Station	(1)	(2)	(3)	(4)	(5)	(6)	(7)
HH	Harrington Harbor *	−0.34	−0.43	−1.4	0.09	1.06	–	–
SJ	St. Johns, Nfld.	1.37	–	0.4	–	0.97	–	−0.13
CH	Charlottetown, P.E.I.	3.08	1.14	0.6	1.94	2.48	0.84	1.38
H	Halifax, N.S.	3.67	2.83	1.6	0.84	2.07	−0.26	0.97
SS	St. John, N.B.	3.09	1.19	0.9	1.90	2.19	0.80	1.09
E	Eastport, Maine	2.70	1.19	0.9	1.51	1.80	0.41	0.70
BH	Bar Harbor, Maine	2.70	1.19	0.8	1.51	1.90	0.41	0.80
PL	Portland, Maine	2.20	0.92	0.7	1.28	1.50	0.18	0.40
PO	Portsmouth, N.H.	1.80	1.49	0.9	0.31	0.90	−0.79	−0.20
BO	Boston, Mass.	2.90	1.49	1.2	1.41	1.70	0.31	0.60
CCC	Cape Cod Canal, Mass.	2.01	1.75	1.7	0.26	0.31	−0.84	−0.79
BB	Buzzards Bay, Mass. *	0.52	1.75	1.7	−1.23	−1.18	–	–
WH	Wood's Hole, Mass.	2.70	1.75	1.7	0.97	1.02	−0.13	−0.08
PR	Providence, R.I.	1.80	1.35	1.5	0.45	0.30	−0.65	−0.80
N	Newport, R.I.	2.70	1.35	1.6	1.35	1.10	0.25	0
NL	New London, Conn.	2.10	1.35	1.5	0.75	0.60	−0.35	−0.50
BR	Bridgeport, Conn.	2.10	1.35	1.5	0.75	0.60	−0.35	−0.50
NR	New Rochelle, N.Y.	2.05	1.35	–	0.70	–	−0.40	–
MO	Montauk, N.Y.	1.90	1.78	1.6	0.12	0.30	−0.98	−0.80
PJ	Port Jefferson, N.Y.	2.70	1.78	1.6	0.92	1.10	−0.18	0
WP	Willets Pt., N.Y.	2.40	1.78	1.6	0.62	0.80	−0.48	−0.30
NYC	New York City, N.Y.	2.70	2.17	1.6	0.53	1.10	−0.57	0
SH	Sandy Hook, N.J.	4.10	1.87	1.7	2.23	2.40	1.13	1.30
AC	Atlantic City, N.J.	3.90	1.87	2.1	2.03	1.80	1.10	0.70
P	Philadelphia, Penn.	2.60	–	1.8	–	0.80	–	−0.30
L	Lewes, Del.	3.10	2.35	2.1	0.75	1.00	−0.35	−0.10
B	Baltimore, Md.	3.20	1.81	1.8	1.39	1.40	0.29	0.30
A	Annapolis, Md.	3.60	1.81	1.8	1.79	1.80	0.69	0.70
W	Washington, D.C.	3.20	1.81	1.8	1.39	1.40	0.29	0.30
SI	Solomon's Is., Va.	3.30	1.81	1.9	1.49	1.40	0.39	0.30
GP	Gloucester Point, Va.	3.38	1.20	1.8	2.18	1.58	1.08	0.48
K	Kiptopeke Beach, Va.	3.10	1.20	1.8	1.9	1.30	0.80	0.20
HA	Hampton Rds, Va.	4.30	1.20	1.8	3.1	2.50	2.00	1.40
P	Portsmouth, Va.	3.70	1.20	1.7	2.5	2.00	1.40	0.90
WI	Wilmington, N.C.	1.80	1.23	1.2	0.57	0.60	−0.53	−0.50
CHN	Charleston, S.C.	3.40	1.01	0.8	2.39	2.60	1.29	1.50
SA	Savannah, Ga.	3.00	0.43	0.6	2.57	2.40	1.47	1.30
F	Fernandina, Fla.	1.90	1.86	0.4	0.04	1.50	−1.06	0.40
MA	Mayport, Fla.	2.20	1.86	0.3	0.34	1.90	−0.76	0.80
DB	Daytona Beach	2.01	1.86	–	0.15	–	−0.95	–
MB	Miami Beach, Fla.	2.30	0.69	0	1.61	2.30	0.51	1.20
KW	Key West, Fla.	2.20	0.69	−0.1	1.51	2.30	0.41	1.20
Average *		2.72	1.50	1.29	1.26	1.47	0.17	0.37
Std. dev.		0.71	0.47	0.60	0.78	068	0.79	0.68
No of stations		(40)	(38)	(38)	(38)	(38)	(38)	(38)

(1) Uncorrected tide gauge data (Lyles et al., 1987 and PSMSL).
(2) Long-term glacio-isostatic corrections; ^{14}C data.
(3) Long-term glacio-isostatic corrections (Braatz and Aubrey, 1987; Peltier, 1986).
(4) Corrected sea level trend; ^{14}C data.
(5) Corrected sea level trend (Braatz and Aubrey, 1987; Peltier, 1986).
(6) Corrected sea level trend minus est. eustatic trend (1.1); ^{14}C data.
(7) Corrected sea level trend minus est. eustatic trend (1.1); (Braatz and Aubrey, 1987; Peltier, 1986).
* Excluding Harrington Harbor and Buzzards Bay.

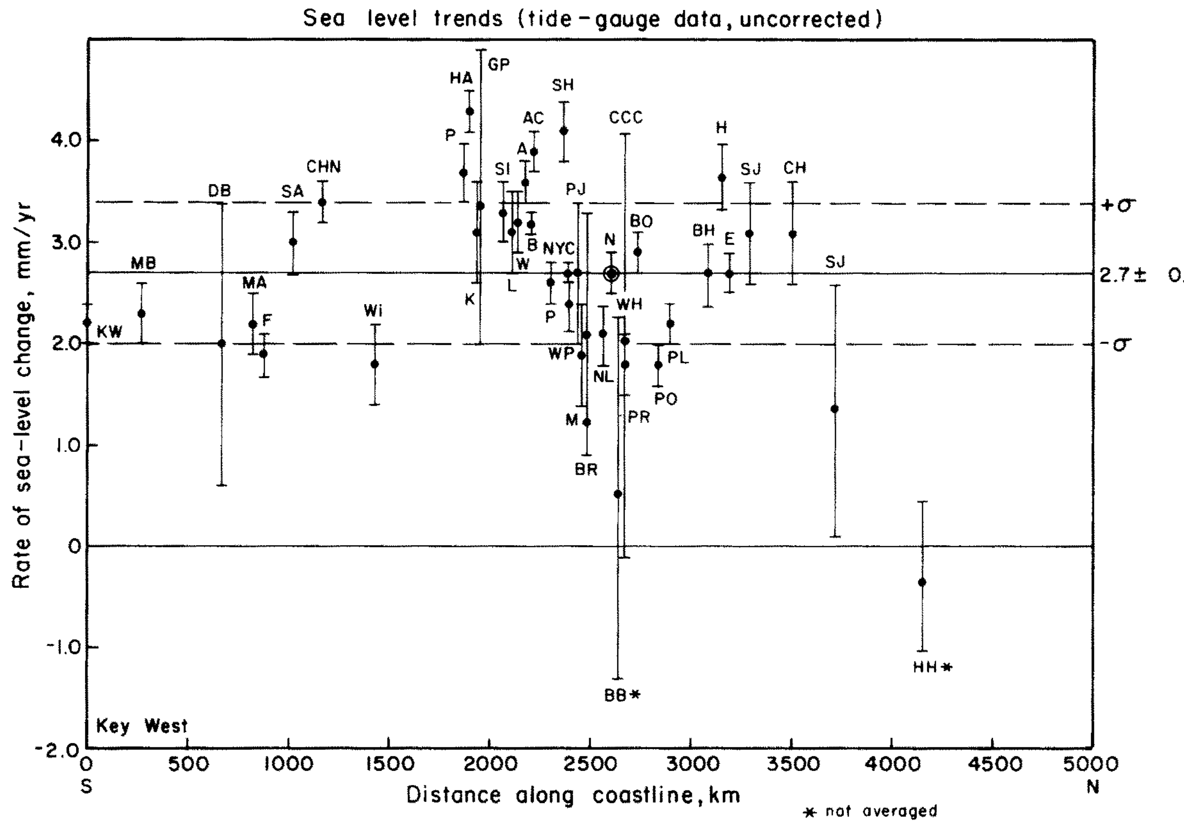

Fig. 7. Present sea level trends (tide-gauge data, uncorrected), after Lyles et al. (1987) and PSMSL. The regional mean is indicated by the solid line; the bounds for plus or minus one standard deviation are shown by a dashed line. The error bars represent the standard error of the slope (see Lyles et al., 1987).

and because of a fairly a short (20 yrs) and very noisy record, has been excluded from the regional mean.

Sea level trends increase between Portland, Maine and Halifax, Nova Scotia, then decrease north of Charlottetown, Prince Edward Island (Fig. 7). The present SL trend for Eastport, Maine (2.7 mm/yr) is close to the regional mean. Reanalysis of leveling observations (Reilinger, 1987) has not reconfirmed the unusually high rates of subsidence reported by Anderson et al. (1984). Instead, the corrected rates of around 1–2 mm/yr are compatible with post-glacial subsidence rates (Table 1, cols. 2, 3, and Belknap et al., 1987), and with the Eastport tide-gauge rate, after subtracting the estimated "eustatic" trend. On the other hand, other historical data, indicating differential subsidence (Anderson et al., 1984), and recent seismicity (Lee, 1985) suggest some possible ongoing neotectonic activity in eastern Maine.

The relatively high SL trends of the Maritime Provinces, Canada (3.1–3.7 mm/yr, Table 1) indicate ongoing post-glacial submergence. Decreasing SL trends toward the north (e.g. St. Johns, Newfoundland, 1.37 mm/yr; Harrington Harbor, −0.34 mm/yr; Table 1, col. 1) illustrate the transition toward a region of glacioisostatic rebound.

Comparison between late Holocene sea level trends based on ^{14}C-dated materials and viscoelastic earth models

The late Holocene trends from ^{14}C and model calculations (Peltier, 1986; Braatz and Aubrey, 1987) are listed in cols. 2 and 3 respectively of Table 1, and Fig. 8. Rates of sea level change derived from both methods agree to within 0.4 mm/yr, in 24 out of 36 paired stations. However, Peltier calibrates his model, using ^{14}C dates, and therefore reasonably good agreement between the two methods can be expected. Since the region is

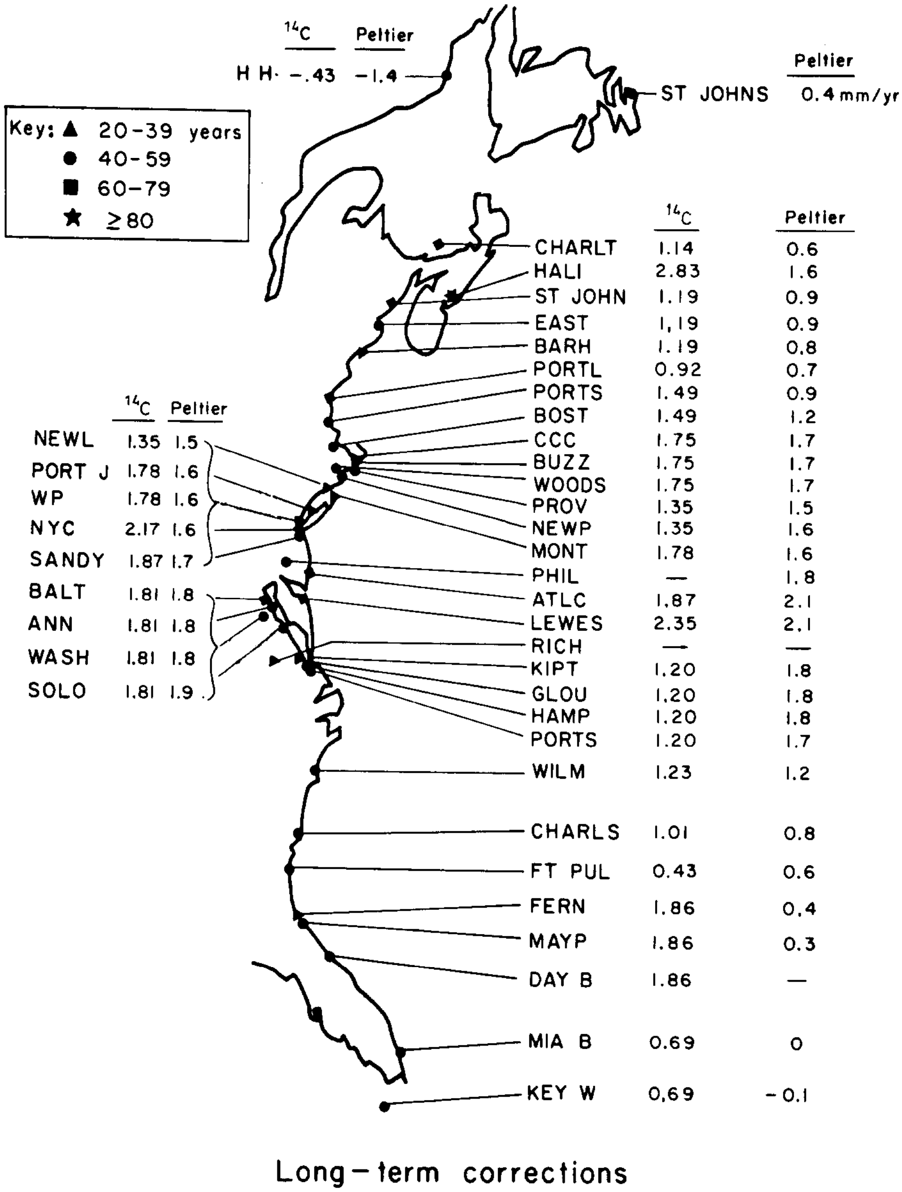

Fig. 8. Map of East Coast tide-gauge stations, comparing long-range sea level trends derived from ^{14}C data and Peltier (1986).

relatively geologically quiescent, much of the late Holocene sea level change can probably be attributed to residual glacioisostatic movements. Nonetheless, the model appears to under-estimate crustal movements in Florida, predicting a nearly zero trend. The ^{14}C-derived trend of 1.86 ± 0.57 mm/yr (standard error of slope) for northern Florida may be too high, because a linear trend may not be strictly valid, due to the wide gap between recent and older data points (Fig. 9A). However, the curve for southern Florida for dates less than 4000 yrs closely approximates a straight line defined by numerous closely-spaced points, with slope 0.69 ± 0.20 mm/yr (Fig. 9B). If errors in the model or in the data base can be discounted, the disparity could reflect neotectonic activity centered on the Peninsular arch (Opdyke et al. 1984). Uplift in northeast Florida may be coupled with gentle offshore tilting, producing apparent late Holocene subsidence, possibly con-

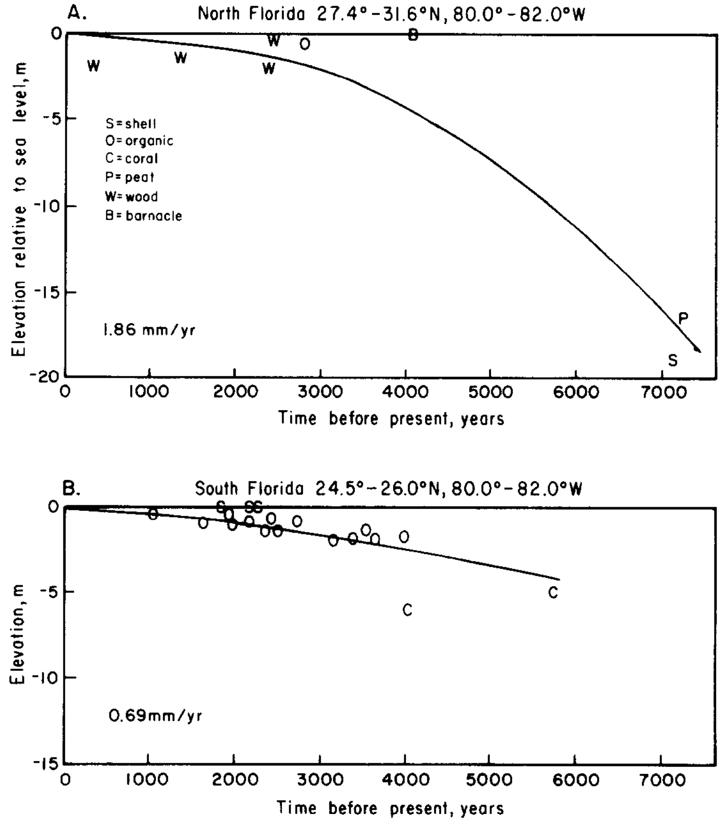

Fig. 9. A. Late Holocene sea level curve for North Florida. B. Late Holocene sea level curve for South Florida.

tinuing into the present, as suggested by geodetic surveys (Brown, 1978; Jurkowski and Reilinger, 1981).

The trend derived from the ^{14}C-based SL curve for southern Chesapeake Bay is distinctly lower (1.2 ± 0.2 mm/yr) than for the northern or upper bay (1.8 ± 0.8 mm/yr) whereas the viscoelastic model predicts the same trend for the whole area (Fig. 10). A lower rate of late Holocene SL rise for the southern Bay is consistent with indications of localized uplift near the entrance to Chesapeake Bay (Harrison et al., 1965; De Alteris and Byrne, 1975). Some suggestion of differential subsidence is also seen in the regional paleosealevel contours (Fig. 4).

The ^{14}C-derived trend for New York City (2.17 ± 0.30 mm/yr) is somewhat higher than the model prediction (1.6 mm/yr); however, the curve is well-constrained by numerous data points which fall on a straight line for dates less than 5000 yrs (Fig. 11). Model and ^{14}C-derived trends are generally in good agreement for New England (Table 1, Fig. 8). The trends obtained for Maine in this study (~ 1.2 mm/yr) agree well with the late Holocene rate of 1.22 mm/yr reported by Belknap et al. (1987).

Cells centered along the upper Bay of Fundy and western Nova Scotia (for which, however, we lack tide-gauge records) show late Holocene trends ≈ 2 mm/yr, independent of the particular cell boundaries selected. No systematic difference was observed between the northern and southern shores of the Bay of Fundy (Fig. 12), thus ruling out detectable renewed vertical movement along Jurassic–Triassic rift structures (Ballard and Uchupi, 1975; Lee, 1985).

Corrected sea level trends

Late Holocene trends from ^{14}C-dating and model calculations (Table 1, cols. 2, 3) have been subtracted from the tide-gauge data. Resulting corrected sea level trends are listed in Table 1, cols 4, 5, and Fig. 13. The regional mean corrected sea level trend (38 stations) is 1.26 ± 0.78 (± 1 σ; ^{14}C

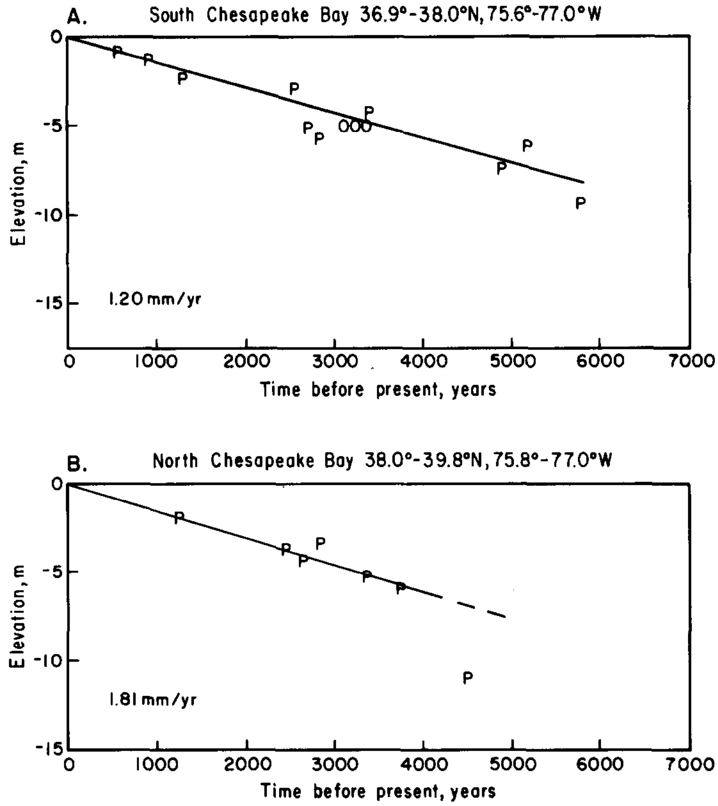

Fig. 10. A. Late Holocene sea level curve for South Chesapeake Bay. B. Late Holocene sea level curve for North Chesapeake Bay.

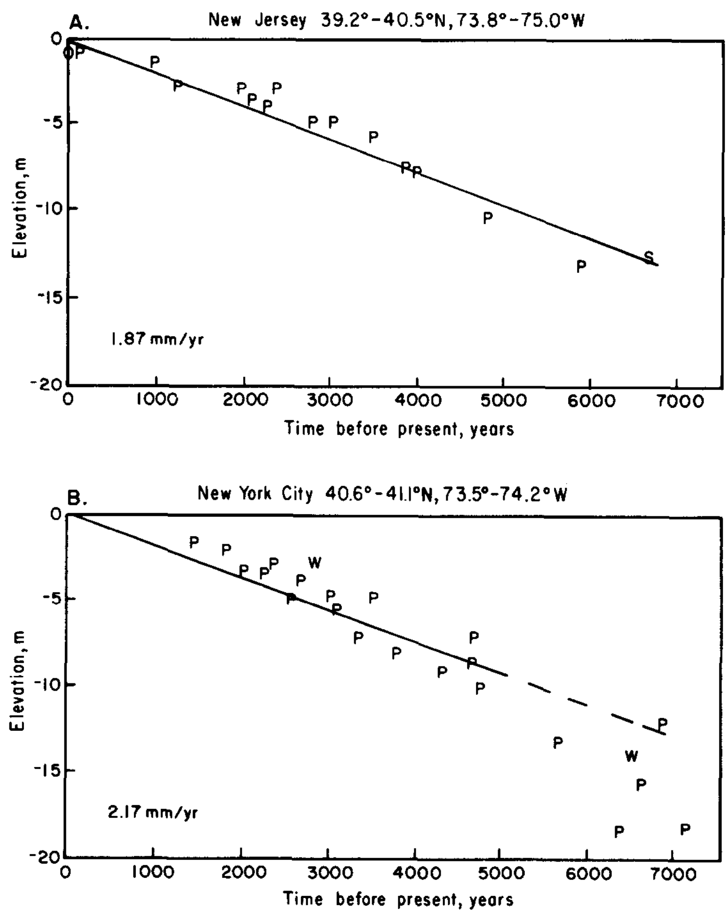

Fig. 11. A. Late Holocene sea level curve for New Jersey. B. Late Holocene sea level curve for New York City area.

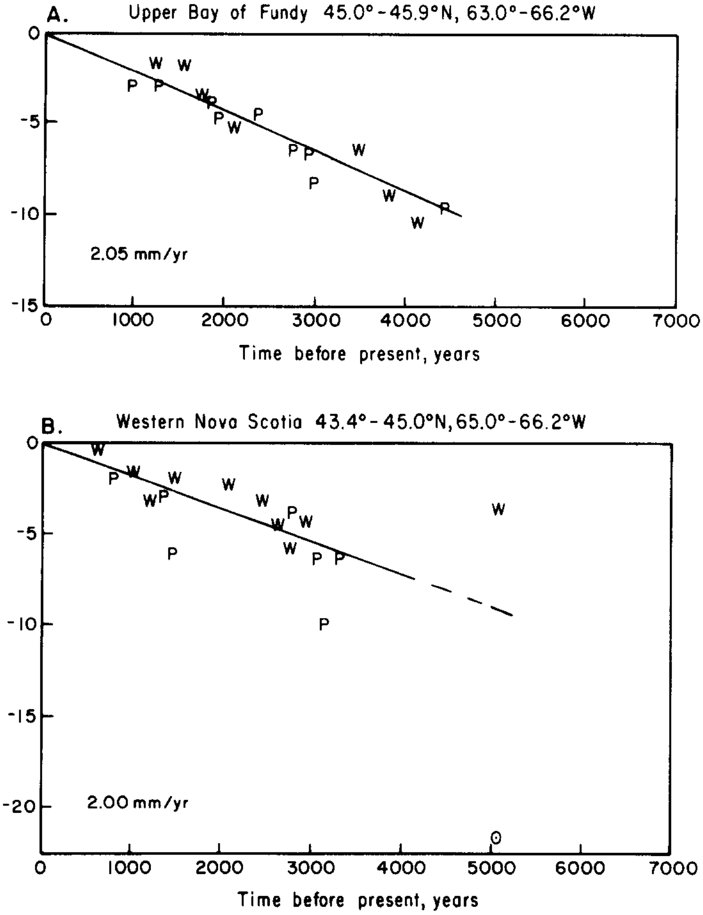

Fig. 12. A. Late Holocene sea level curve for Upper Bay of Fundy. B. Late Holocene sea level curve for southwestern Nova Scotia.

data), and 1.47 ± 0.68 mm/yr (model), respectively. As can be seen from the data, presently the sea level is rising around two times faster than the long-term late Holocene trend along the U.S. East Coast. The estimated global eustatic trend of 1.1 mm/yr (taken as the mean of two different averaging methods—Gornitz and Lebedeff, 1987) is subtracted and the residuals listed in cols. 6 and 7 of Table 1. The regional residual means of 0.17 ± 0.79 and 0.37 ± 0.68 mm/yr, respectively, are small, suggesting that most of the subsidence can be attributed to eustatic and glacioisostatic processes. However, the non-zero value and high spatial variability imply the contribution of a small neotectonic component, as well.

Stations with ^{14}C residual trends lying outside ± 1 σ of the regional mean include: Daytona Beach–Fernandina (north Florida); Savannah–Charleston; Portsmouth (Virginia)–Hampton Roads (Virginia); Atlantic City–Sandy Hook; Montauk (New York); Providence (Rhode Island); Cape Cod Canal (Massachusetts); and Portsmouth (New Hampshire). The northern Florida stations have lower than average residuals, because of a possibly erroneously high late Holocene trend (see above). The high residual at Charleston may be largely caused by groundwater withdrawal (Poley and Talwani, 1986); however, given the historic seismicity, neotectonic subsidence remains a possibility. Late Holocene sea level fluctuations in the Charleston area may reflect neotectonic movements (Colquhoun and Brooks, 1987). Southern Chesapeake Bay shows recent apparent high subsidence, in contrast to possible late Pleistocene–Holocene uplift (Harrison et al., 1965; De Alteris and Byrne, 1975). However, some of the anomalous subsidence near Norfolk (Virginia) may be produced by groundwater withdrawal (Davis, 1987). Montauk, Providence and Cape Cod Canal have below average residuals because of slightly below average rates of recent sea level rise, combined with relatively higher late Holocene trends, suggesting the possibility of uplift. Any possible renewed movement along offshore Mesozoic rift faults on the Long Island Platform (Hutchinson et al., 1986) affecting these three stations is not readily apparent, since they do not lie close to any of the mapped faults. On the other hand, the substantial difference in the corrected RSL rates between two adjacent stations on Long Island: Montauk (0.12 mm/yr) vs. Port Jefferson (0.92 mm/yr) suggests a relative uplift of eastern Long Island, which is supported by geomorphology (emergent beach ridges and tombolo gravels on the eastern tip of Long Island —Fairbridge, 1989). The observed sense of displacement is consistent with possible motion along the New York Bight Fault, south of central Long Island, which may have still been active during the Quaternary (Hutchinson and Grow, 1985).

Discussion and conclusions

The hingeline between glacial isostatic subsidence and uplift has shifted north by at least 5° latitude, from around 43° to 48° N, over the last 14,000 yrs (Fig. 3a–n). The zone of maximal subsidence has similarly moved north from around 36° to 42° N prior to 11,000 yrs to 44°–46° N during the last 5000 yrs. The latitudinal projections of paleosealevel indicators clearly delineate the northward migration of the collapsed peripheral bulge (see also Fairbridge and Newman, 1968; Walcott, 1972; Clark et al., 1978; Peltier, 1986; Pardi and Newman, 1987).

The sequence of shoreline profiles shown in Fig. 3 differs from Clark (1981) who predicted a downward tilt from Florida to New Jersey, between 4000–11,000 yrs B.P., a flat profile around 11,000 yrs B.P., and a reversal of the tilt (toward the south) for older sediments. Instead, the downward tilt to the north between Florida and New Jersey increases with increasing age, and is defi-

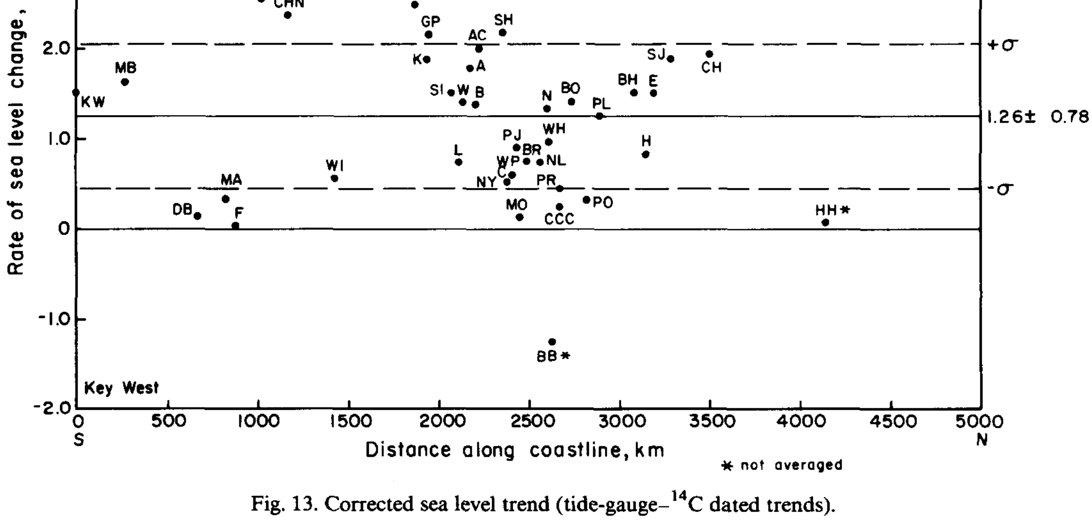

Fig. 13. Corrected sea level trend (tide-gauge–^{14}C dated trends).

nitely not flat ~ 11,000–12,000 yrs B.P. The data also refute Blackwelder's (1980) claim of undeformed shorelines between 9000–12,000 yrs B.P. Furthermore, although paleosealevels are deeper at the maximal peripheral bulge collapse off the New York-New Jersey coast (southernmost extent of the ice sheets), no discrete "inflection points" (Dillon and Oldale, 1978) are observed. Estuaries and bays such as the upper Chesapeake and Delaware Bays, and the upper Bay of Fundy–Nova Scotia have subsided more than adjacent areas within the past 7500 yrs (Figs. 4 and 6).

Recent sea level data for the East Coast show annual increments that range between 2–4 mm (Table 1, col. 1). One major component is eliminated (this paper), namely a long-wavelength, long-period trend. As discussed above, this is largely glacioisostatic (Fig. 8), but may also include flexure from slower-changing loads, and sediments. Subtraction of the long-range trend reduces the regional mean from 2.72 ± 0.71 mm/yr to 1.26 ± 0.78, thus accounting for about half of the signal. The corrected regional mean trend lies within a few tenths of a mm/yr of the estimated global eustatic sea level rise (taken as 1.1 mm/yr; Table 1). After subtraction of long-term crustal and eustatic components, the residuals still show significant spatial variability. The intraregional variability comes from yet other factors, of which four possibilities are now considered: (1) river runoff, (2) sediment loading and lithosphere cooling (3) groudwater pumping, and (4) neotectonic activity. While changes in ocean circulation affect sea level, these changes tend to be coherent over long wavelengths (Komar and Enfield, 1987), and therefore are not likely to affect the short-wavelength neotectonic signal.

Meade and Emery (1971) have suggested that 7–21% of the total variation in mean annual sea level along the East Coast could be caused by fluctuations in river runoff, which, in turn, is related to rainfall.

Flexure from sediment loading and lithospheric cooling is slow and is a generally insignificant cause of differential sea level change regionally, over a 100 to 10,000 year period. The average total subsidence rate, from these processes along the U.S. East Coast, has remained fairly constant, around 0.03 mm/yr for the last 135 m.y. (Heller et al., 1982). Holocene subsidence rates over the entire continental shelf due to these processes, are not likely to differ substantially from the Cenozoic rate.

On the other hand, sedimentation and compaction could be locally important, near major rivers, in particular the Chesapeake–Delaware drainage basins, of the eastern U.S. (Curtis et al. 1973). Suspended sediment loads are affected by such factors as source area bedrock geology, land-use practices (forest clearing, farming, etc.), and rainfall. However, if present trends are at least qualitatively representative of the late Holocene, the Chesapeake and Delaware Bays may have accumulated thicker piles of sediments than adjacent river basins, inducing a greater amount of subsidence in the estuary, as opposed to the near shore environment (Fig. 4 and Meade, 1982).

Withdrawal of subsurface fluids (oil, gas, water) has caused subsidence and exacerbated the natural sea level rise at a number of coastal sites such as Savannah, Georgia (Davis, 1987). However, the tide-gauge at Fort Pulaski lies beyond the area of subsidence at Savannah and therefore it does not affect our results (Davis et al., 1977, Davis, 1987). Other cities on the Coastal Plain, such as Charleston, are also potentially subject to this problem (Poley and Talwani, 1986). The magnitude of this effect can be estimated by comparing sea level data for stations along the coastal plain from Key West, north to Long Island with data for localities on crystalline bedrock, from New England north to St. John's, Newfoundland. The mean rate of sea level rise for 24 tide stations along the Coastal Plains is 2.90 ± 0.74 mm/yr (σ) and for 16 on crystalline rock is 2.46 ± 0.59 mm/yr. Although the average on bedrock is slightly lower, the difference between these two means is statistically insignificant at the 95% confidence level (student's t-test). However, the regional average may mask large local effects from water withdrawal, at some stations.

This leaves neotectonic movement as one of the most likely factors contributing to intraregional variance in sea levels, after accounting for isostatic loading (Brown, 1978; Braatz and Aubrey, 1987). The sea level data presented above also support this view.

The topography of the Coastal Plain provides additional support for gentle ongoing coastal deformation, particularly in the vicinity of Chesapeake Bay and the Cape Fear arch. The marked difference in elevation between both sides of Chesapeake Bay may have been caused by differential erosion, or uplift. The latter possibility is more likely, given gentle regional oscillatory movements during the Neogene ("dancing basins and arches", Newell and Rader, 1982). Middle Miocene sediments thicken northeast toward the Salisbury Embayment (Northern Chesapeake Bay), whereas late Miocene to Pliocene units thicken southeast toward the Albermarle Embayment (Southern Chesapeake Bay; Newell and Rader, 1982), indicating a shift in the locus of subsidence. By the Pleistocene, evidence from river terraces on the west bank of Chesapeake Bay suggests renewed tilting toward the east (Newell, 1985). Holocene ^{14}C-dated materials (Fig. 4) suggest a continued northeast tilt toward the Salisbury Embayment and concurrent uplift in Southern Chesapeake Bay (Albermarle Embayment; Harrison et al. 1965; De Alteris and Byrne, 1975). In historic times, the subsidence pattern has shifted again to the Delaware Bay, and the Albermarle Embayment, near Norfolk, Virginia (Holdahl and Morrison, 1974). Vertical movements in the latter area are also consistent with tide-gauge records (Fig. 10). This subsidence may, in part, be anthropogenically-induced (Davis, 1987). By contrast, the Cape Fear arch has remained continually high during the Cenozoic (Colquhoun and Brooks, 1987).

Uchupi and Aubrey (1988) have interpreted observed variations in sea level along the coast in terms of suspect terranes that were accreted to the eastern edge of the North American craton during the Paleozoic. However, their sea level trends, corrected for glacioisostasy, are weakly correlated with the proposed terranes, and instead are more closely related to Mesozoic–Cenozoic structures (see Braatz and Aubrey, 1987).

A major finding of this study is the absence in eastern North America of zones that show marked differential vertical movement indicative of active fault displacement, in spite of some studies that suggest high shear rates (e.g., Zoback et al., 1985). Sea level data should be able to resolve relative differences in vertical components of displacement over 0.5 mm/yr for the Holocene. Differential movements exceeding such rates over geologic periods, should result in pronounced morphologic expression. The sea level data are consistent with geologic evidence for the absence of large faults with substantial Neogene displacement (with some exceptions; see for example Newman et al., 1987).

The absence of prominent active faults does not preclude the occurrence of large earthquakes, such as the one in Charleston, South Carolina in 1886. The remarkable juxtaposition of substantial seismicity with absence of active faults suggests that deformation may be spread over a wide area and that the seismicity in any one area is clustered in time (Coppersmith, 1988).

High rates of deformation given by geodetic data (e.g. Holdahl and Morrison, 1974; Brown, 1978; Poley and Talwani, 1986), are frequently inconsistent with geologic, geomorphic and seismic data. Geodetic results present a non-unique tectonic interpretation, either because the data may be contaminated by systematic errors (Reilinger, 1987) or because the deformation can be attributed to other factors, such as groundwater withdrawal (Poley and Talwani, 1986). In some places, (i.e. Charleston), sea level anomalies are colocated with geodetic anomalies. While such coincidences help reinforce the credibility of the results, they cannot always constrain the causes.

Zoback et al. (1985) propose a high rate of right shear along the lower Hudson Valley, based on a reexamination of triangulation data. Because current rate of moment release in the region is grossly insufficient to account for this strain, they suggest that the deformation is primarily elastic, which will eventually lead to a major earthquake. Sea level data do not show a prominent anomaly in the New York City area. While sea level is insensitive to horizontal shear, the high rate of strain proposed by Zoback et al. should produce a vertical component of defomation. The absence of a sea level anomaly is inconsistent with a major active structure along the southern Hudson Valley. Farther north along the Hudson River, on the other hand, some differential movement along the Ramapo border fault zone of the Newark basin, in northern New Jersey, and adjacent New York, has been deduced from paleosealevel data (see Figs. 1 and 4 in Newman et al., 1987).

However, the sea level data do indicate several anomalous areas, suggestive of gentle epeirogenic movements. While further field verification needs to be obtained, the following anomalous stations or areas are considered potential candidates for further study of possible vertical movements, especially in light of known historic or Neogene activity:

(1) Subsidence between Savannah (Georgia), Charleston (South Carolina)—in part anthropogenic, in part neotectonic (Lyttle et al. 1979; Poley and Talwani, 1986).

(2) Uplift along the Cape Fear Arch (Markewich, 1985).

(3) Southern Chesapeake Bay—late Pleistocene to Holocene uplift (Harrison et al., 1965; De Alteris and Byrne, 1975), followed by more recent subsidence, in part anthropogenic (Holdahl and Morrison, 1974; these data).

(4) Montauk; possible uplift of eastern Long Island (Fairbridge, 1989), related to renewed movement along the New York Bight Fault south of Long Island (Hutchinson and Grow, 1985).

(5) Other anomalous stations near Long Island Platform: Providence (Rhode Island), Cape Cod Canals (Massachussetts).

Acknowledgments

This research was carried out under NASA Cooperative Agreement NCC 5-29, Task A and U.S. Nuclear Regulatory Commission NRC04-85-111-02. The late Professor W. Newman, Queens College, New York generously gave permission to use his data base for this study. We thank Dr. D.T. Pugh, former Director, Permanent Service for Mean Sea Level, Bidston Observatory, Birkenhead, England, for sending us a computer tape of tide-gauge records; and Dr. S. Hicks, National Ocean Service, National Oceanic and Atmospheric Administration, Rockville, Maryland for providing updated tide-gauge records for the United States. We benefitted from interesting discussions with Dr. S. Lebedeff, Goddard Institute for Space Studies (GISS) and Prof. R.W. Fairbridge, Columbia University and GISS.

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